VERTICAL WIND SHEARS NEAR THE CORE OF THE JET STREAM OVER THE NORTHEASTERN UNITED STATES, AUGUST 1–2, 1958

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1. INTRODUCTION

One of the most valuable uses of knowledge of vertical wind shears is their application to planning optimum flight paths for jet aircraft. In forecasting winds for aircraft, their use as an incremental term has tended to simplify the task of interpolating winds between standard constant pressure levels. Furthermore, from the standpoint of flying safety and passenger comfort, the observation of extreme vertical shears has been helpful in identifying regions of possible clear air turbulence.

On August 1, 1958 there was evidence that strong vertical wind shears existed near the core of the jet stream over the northeastern United States, and within 24 hours the magnitude of these shears had diminished considerably. Wind shears computed directly from rawin reports (see table 1) indicated values as large as 21 kt./1000 ft. Shears of this magnitude greatly exceed the value of 10 kt./1000 ft. which has been regarded as critical enough to divert a jet aircraft from its optimum flight path [1]. The purpose of this article is to analyze the conditions under which these extreme vertical wind shears occurred and to demonstrate their effective usefulness by integrating them into a vertical wind shear chart.

2. RELATIONSHIP OF SYNOPTIC FEATURES AND JET STREAM MODEL

At 0000 GMT, August 1, the 250-mb. chart (fig. 1A), which usually represents the standard level nearest the core of the jet stream, indicated a Low over eastern Canada with a flat short-wave trough extending southwestward from the Gulf of St. Lawrence to southeastern Michigan. The wind flow over the northeastern United States was generally from the west or northwest and contained two jet streams which tended to approach each other over New England. One wind speed maximum was located near Maine with speeds of at least 120 kt., and the other was near Albany, N. Y. with speeds of more than 100 kt.

Table 1.—*Occurrence of large vertical wind shears, August 1–2, 1958*

Location	Date	Time (GMT)	Magnitude (kt./1000 ft.)	Height (Thsds. of ft.)	Depth (ft.)
Caribou, Maine	1	0000	−10	35–40	5000
Idlewild, N. Y.	1	0000	20	40–41	1000
Idlewild, N. Y.	1	0000	−12	41–43	2000
Dayton, Ohio	1	0600	−9	44–46	2000
Caribou, Maine	1	1200	−18	46–48	2000
Dayton, Ohio	1	1800	15	28–30	2000
Dayton, Ohio	1	1800	−9	30–32	2000
Pittsburgh, Pa.	2	0000	−21	40–43	3000
Idlewild, N. Y.	2	0000	9	30–35	5000
Dayton, Ohio	2	1200	9	32–40	8000

Both jet stream positions seemed reasonably well located with respect to horizontal temperature gradient and contour spacing. Associated with the northern jet was a steep tropopause slope, while the southern jet appeared to be related to the surface frontal system. The surface front at this time was quasi-stationary with weak minor waves and extended westward from the Atlantic Ocean to a position south of Idlewild, on Long Island, N. Y., and into central Ohio. In view of the upper confluent pattern superimposed on the surface front, it seemed unlikely that any strong development was possible at this time.

In conjunction with this synoptic pattern, Caribou, Maine reported a maximum vertical wind shear near the core of the jet of −10 kt./1000 ft. and Idlewild reported 20 kt./1000 ft. The strong shear at Caribou occurred over a 5000-ft. layer located above the jet stream, to the rear of the trough, and in relation to a steep tropopause slope. When compared to the jet stream models described by either Endlich [2] or Riehl et al. [3], the magnitude of Caribou's vertical wind shear seemed quite valid, especially since it was apparently associated with strong horizontal wind shears to the left of the jet.

While the northern jet stream was either over or just south of Caribou, the southern jet was at least 120 miles north of Idlewild. As would be expected, there was no evidence of strong horizontal temperature gradients south of the jet stream, yet Idlewild reported a positive shear of 20 kt./1000 ft. over a 1000-ft. layer beneath the peak wind and a shear of −12 kt./1000 ft. over a 2000-ft. layer above the peak wind. The short-wave trough at this time was about 250 miles northwest of Idlewild and the slope of the tropopause in the area was negligible.

According to the jet stream model, the largest vertical wind shears are usually found just above the core of the jet stream and on the left or cold side in conjunction with strong horizontal wind shears and temperature gradients. Strong vertical shears have also shown some tendency to be associated with strong peak winds, but the correlation was found by Dvoskin and Sissenwine [4] to be rather low. When compared to the jet stream model, the magnitude of Idlewild's extreme vertical shears seemed highly questionable.

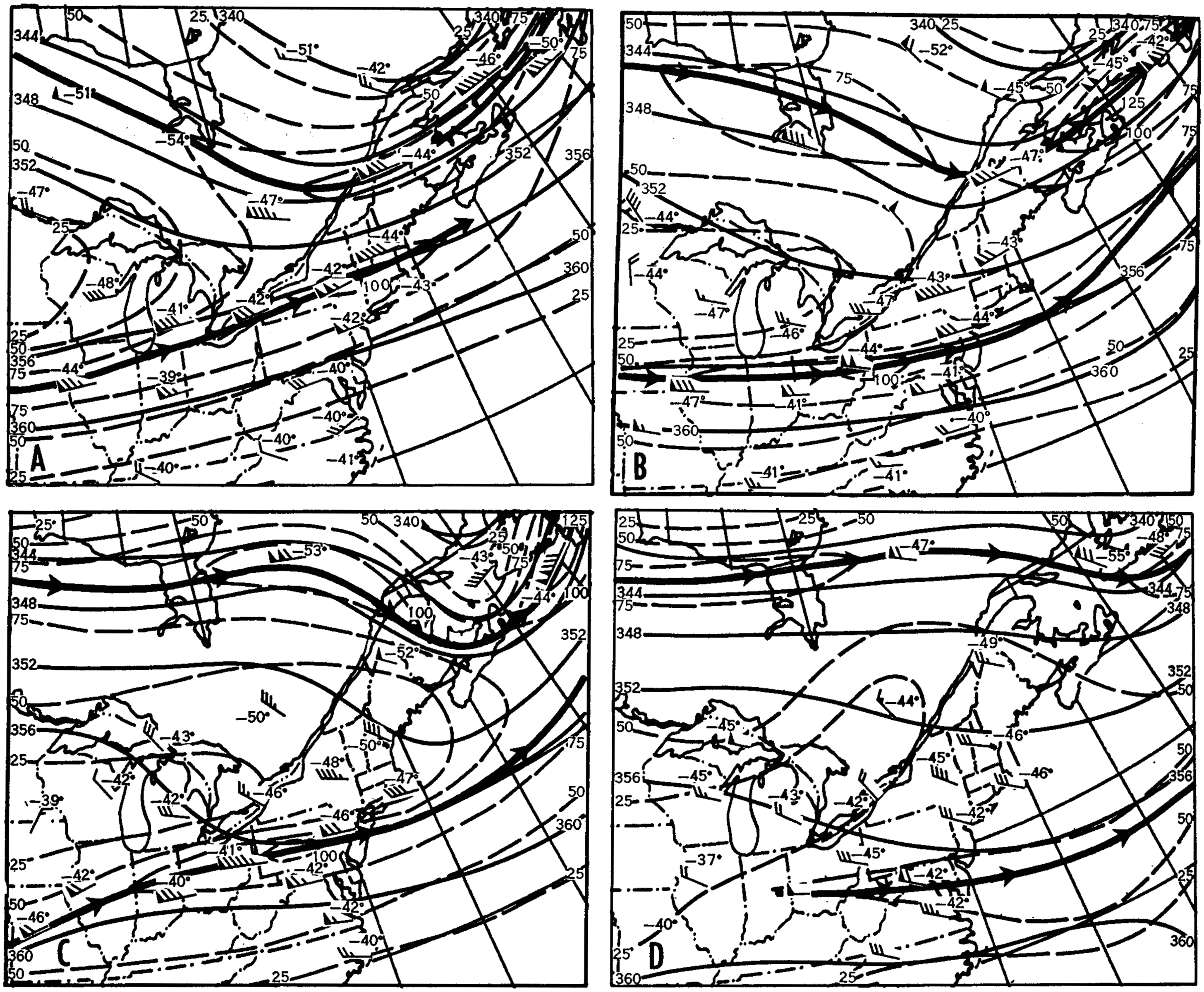


FIGURE 1.—250-mb. contour pattern including jet streams and isotach analysis (dashed). (A) 0000 GMT, August 1; (B) 1200 GMT, August 1; (C) 0000 GMT, August 2; and (D) 1200 GMT, August 2, 1958. The eastward progression of the flat short-wave trough is indicated with its influence on the southern jet stream. The strong horizontal wind shears appeared in conjunction with the largest horizontal temperature gradient.

Although the horizontal temperature gradient was rather weak, a computation was made to determine a reasonable vertical wind shear value for Idlewild, using the thermal wind equation

$$\frac{\partial V}{\partial z}=\frac{g}{fT^*}\frac{\partial T^*}{\partial s}$$

where V is the speed of the westerly wind, g is acceleration of gravity, f is Coriolis parameter, T^* is virtual temperature, z is the vertical coordinate, and s is the horizontal coordinate, positive southward. For an average temperature gradient south of the jet of 1° C. per 125 miles at the 250-mb. level, the vertical wind shear at Idlewild was computed to be a little more than 1 kt./1000 ft. Actually, the computation showed no vertical wind shears for the temperature gradient with respect to Albany since the temperatures at Idlewild and Albany were identical. For the temperature gradient with respect to Washington, the shear at Idlewild was computed tc be 2.5 kt./1000 ft.

From the large difference between the theoretical and observed vertical wind shears at Idlewild, it seemed likely that the strong reported shears were related to either instrumental error, reporting procedure, or microscale phenomena. In any case, the shears were regarded as having little or no effective value for prognostic purposes.

At 1200 GMT, August 1, the short-wave trough at the 250-mb. level (fig. 1B) was still quite flat but appeared to be located along a line through Maine, west of Albany, and in the vicinity of Pittsburgh. The northern jet stream was in about the same latitudinal position with respect to Caribou, but the wind maximum had moved on eastward to Nova Scotia. The southern jet stream had moved southward to an east-west line through southern Pennsylvania with a wind maximum near Pittsburgh exceeding 100 kt. The horizontal temperature gradient to the north of the southern jet was estimated at 3° C./180 mi.

Albany's rawin at this time presented an analysis problem in properly locating the jet since its wind at the 250-mb. level was within about 5 kt. of that at Pittsburgh. However, in view of the horizontal temperature gradient north of the jet and the apparent streamline relationship between Buffalo, Albany, and Portland (fig. 1B), Albany's wind was regarded as unrepresentative.

At the surface (fig. 2) the front was still weak with minor waves but had shifted a little southward to a position through Virginia and Kentucky. Except for Maine and portions of northern New York, there was a wide band of cloudiness over the area; and rain prevailed over much of Pennsylvania and nearby States. The position of the rainfall area had the conventional relation to the jet stream position and helped confirm the upper-air analysis.

Coincident with the synoptic features for 1200 GMT, August 1, a vertical wind shear of −18 kt./1000 ft. over a 2000-ft. layer above the peak wind was reported by Caribou. While this value seemed rather excessive in relation to a reported peak wind of only 72 kt., there was some continuity with the past 12 hours for a large shear to exist. In relation to the tropopause, this large shear occurred in a zone of steep slope which was estimated at 75 ft. per mile. With respect to the jet stream, it was difficult to determine from surrounding data whether the jet was north or south of Caribou; but in any event, the horizontal wind shear appeared rather weak on both sides of the jet and implied a smaller vertical wind shear than that reported. Computation of a vertical wind shear at Caribou by the thermal wind equation gave a value of only .86 kt./1000 ft. In view of these features, Caribou's shear of −18 kt./1000 ft. seemed much too large and had little or no useful value.

Along the southern jet stream there were no vertical wind shear values exceeding the critical 10 kt./1000 ft. The largest shears computed directly from rawin data were 8 kt./1000 ft. at both Dayton and Pittsburgh which were near the base of the short-wave trough. Idlewild and Flint, Mich. had wind shear values of 3 kt./1000 ft. and 5 kt./1000 ft., respectively.

Although these shear values did not conflict with the jet stream model to any great extent, it was rather surprising that Pittsburgh's vertical shear was not larger. The core of the jet stream appeared to be located either over or just south of Pittsburgh, the peak wind was 102 kt., and the 250-mb. temperature gradient over a 130-mi. distance north of the jet was estimated at 1° C./25 mi. Nevertheless, a computed vertical wind shear value of only 5.4 kt./1000 ft. was obtained from the thermal wind equation. The slope of the tropopause over Pittsburgh at this time was negligible.

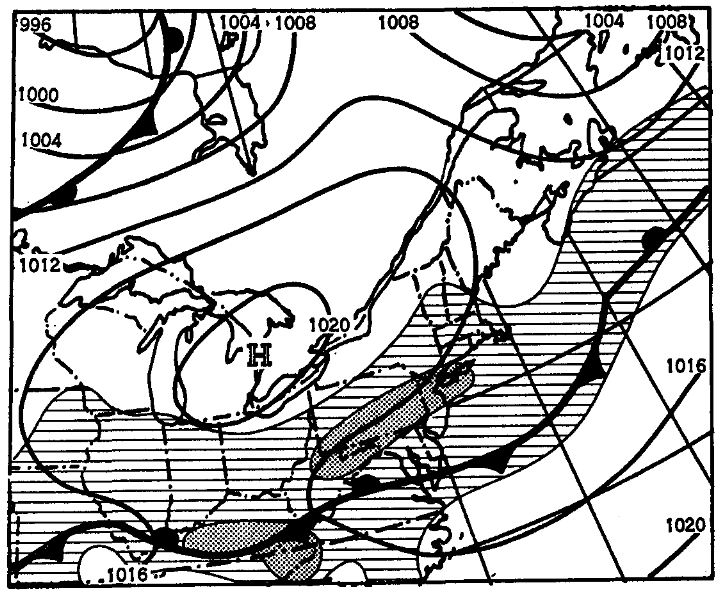


FIGURE 2.—Surface chart for 1200 GMT, August 1, 1958, which was characteristic of the entire period. The surface front and the weather pattern appeared to be related to the southern jet. Area of cloudiness is hatched; precipitation areas are stippled.

At 1800 GMT, August 1, wind reports from Dayton indicated a positive vertical wind shear of 15 kt./1000 ft. over a 2000-ft. layer below the peak wind and −9 kt./1000 ft. over a 2000-ft. layer above the peak wind. Since it was assumed that there had been little or no change in the synoptic pattern since 1200 GMT, it seemed rather unusual for Dayton to have such strong vertical wind shears, expecially in conjunction with a peak wind of only 62 kt. Another suspicious factor was that the peak wind was reported at the 30,000-ft. level when only 6 hours previous it had been at 42,000 ft. From these facts with respect to continuity and the jet stream model, it appeared that these large vertical shears were unrepresentative and had little or no effective value for forecasting.

By 0000 GMT, August 2, the short-wave trough at the 250-mb. level (fig. 1C) appeared to be near the east coast of the United States, but the contours remained cyclonically curved to the north of the southern jet stream from the Midwest to the Atlantic Ocean. The southern jet stream position was along a line north of Dayton, over Pittsburgh, and south of Idlewild, while the northern jet was about 100 mi. northeast of Caribou. The wind flow pattern was still west to northwest, but wind speeds at Albany, Caribou, Portland, and Nantucket had decreased to 50 kt. or less. Winds along the southern jet remained fairly strong with a maximum of at least 120 kt. over southern Pennsylvania.

The tropopause chart at this time showed an unusual ridge across Pennsylvania separating low tropopause heights over Ohio from those in New England. The slope of the tropopause normal to the jet was greatest between Idlewild and Albany with a value estimated at 50 ft. per mile. Parallel to the jet stream, the tropopause slope between Dayton and Pittsburgh was about 34 ft. per mile.

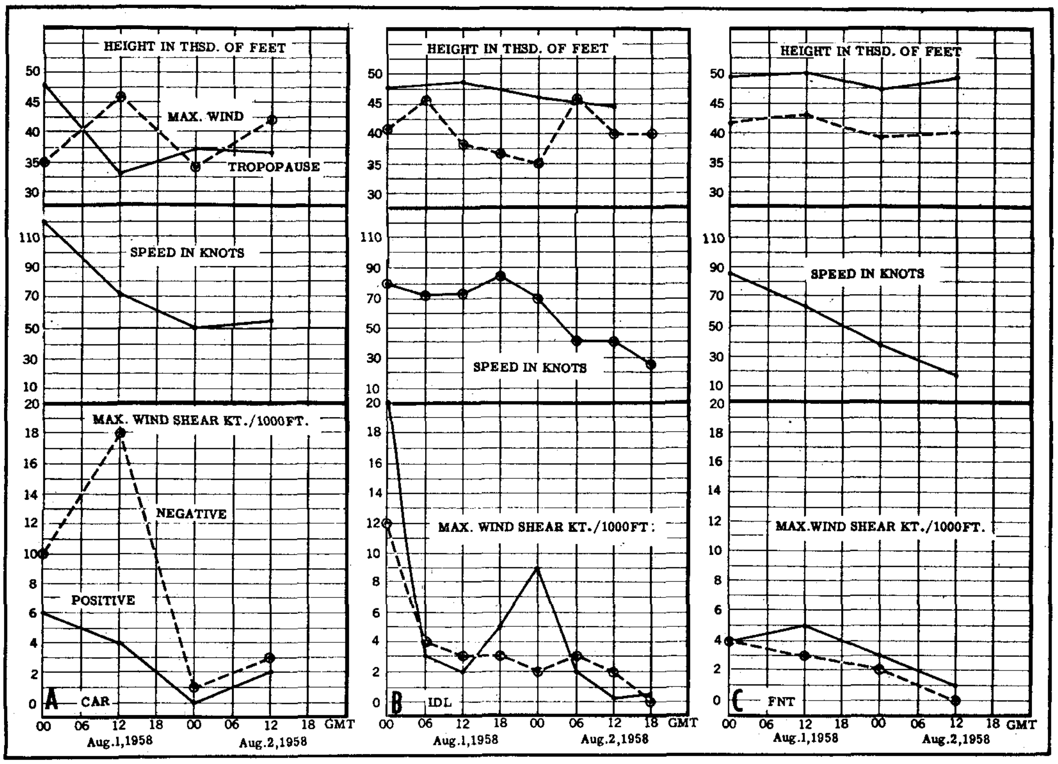


FIGURE 3.—A composite time graph which shows the random fluctuation of several wind field parameters at stations near the core of the jet stream. Upper panel compares the height of tropopause with the height of the peak wind (dashed line). The center panel shows the magnitude of the peak wind; and the bottom panel compares the maximum positive vertical wind shears with the negative (dashed line) wind shears. All wind values were taken directly from reported rawin data. (A) Caribou, (B) Idlewild, (C) Flint.

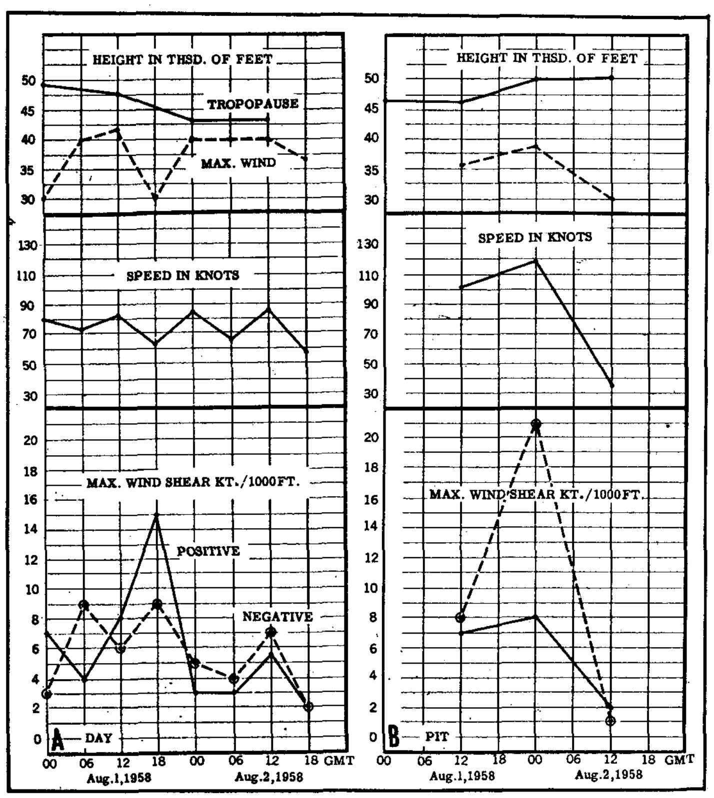


FIGURE 4.—Composite time graph of wind field parameters for (A) Dayton and (B) Pittsburgh.

Concurrent with these synoptic features, Pittsburgh reported a vertical wind shear of −21 kt./1000 ft. over a 3000-ft. layer directly above the peak wind of 119 kt. This strong shear was compatible with both the strength of the peak wind and the isotach pattern, which suggested that the horizontal wind shears were approaching a maximum value. In figure 1C, the horizontal wind shears near Pittsburgh and to the south of the jet stream were estimated to be a little less than 25 kt./deg. lat., and those to the north of the jet at about twice that value. Riehl's [5] computations for maximum horizontal shear for straight or slightly cyclonic flow have shown similar values.

Although the occurrence of this strong shear at Pittsburgh appeared quite reasonable with respect to the jet stream model, the magnitude of the shear was regarded as excessively large. Kochanski [6] has computed vertical shear of this magnitude by using horizontal temperature gradients of 10° C./deg. lat. In this case, the horizontal temperature gradient normal to the jet was only 1° C./26 mi., and the theoretical shear value computed from the thermal wind equation was only 5.4 kt./1000 ft. Although this computed value appeared too low at first, it seemed in good agreement with the value of 6 kt./1000 ft. obtained by averaging Pittsburgh's vertical wind shears over a 10,000-ft. layer above and below the level of maximum wind. Therefore, in view of this analysis, the extreme vertical wind shear at Pittsburgh at 0000 GMT, August 2, was assumed to be the result of either some microscale phenomenon of short duration or of errors in wind measuring equipment or reporting procedure.

At the same time, Pittsburgh had also reported a positive shear of 8 kt./1000 ft. over an 8000-ft. layer beneath the peak wind. This value seemed compatible with the shear of 5 kt./1000 ft. reported at Dayton and 9 kt./1000 ft. at Idlewild. All of these vertical wind shears were in general agreement with the jet stream model and acceptable for a useful analysis. As a matter of interest, a comparison between theoretical wind shears and the above shears determined directly from rawin data showed the latter to be about 3 times larger than those computed by the thermal wind equation.

By 1200 GMT, August 2, the short-wave trough at the 250-mb. level (fig. 1D) was off the east coast of the United States. The southern jet stream was in about the same latitudinal position as 12 hours earlier, but wind speeds had diminished considerably. As might be expected, both positive and negative vertical wind shear values had also diminished. With the exception of Dayton, most wind reports along the jet stream indicated vertical shears of 3 kt./1000 ft. or less. Dayton reported a positive shear of 9 kt./1000 ft. and a negative shear of 7 kt./1000 ft. over 8000-ft. and 10,000-ft. layers, respectively. Although these shears were neither extreme nor in direct conflict with the jet stream model, it seemed curious that they should have occurred with such strength near the end of a declining jet and in an area of rather weak horizontal temperature gradient.

3. COMPATIBILITY OF DATA

Since it has been established that wind data are subject to considerable variation near the core of the jet stream, figures 3 and 4 have been prepared to summarize the fluctuations during August 1–2 and to compare some of the wind field parameters. At 1200 GMT on both August 1 and 2, Caribou (fig. 3A, upper panel) presented an analysis problem when the height of the tropopause appeared to be out of phase with the height of the peak wind. Actually, subsequent investigation disclosed the presence of two tropopause points, and both met the currently adopted definition. Since the jet axis had been analyzed within 150 miles of Caribou (fig. 1B), this out-of-phase relationship appeared to conflict with the jet model of Riehl [3] but seemed compatible with Endlich's model [2].

Endlich has found similar out-of-phase relationships between the height of the tropopause and that of the peak wind and has classified them as type "A" tropopauses that are characteristic of the anticyclonic side of the jet. Type "A" is described as a high cold tropopause in conjunction with a lower, weaker tropopause that is about 10° C. warmer than the higher one. The average height of the lower tropopause was found to be 4000 ft. below the average height of the peak wind. While Caribou's upper-air data exhibited some similarities to Endlich's type "A" tropopause, it seemed quite likely that the analysis problem could best be solved by arbitrarily increasing the wind speed adjacent to the lower tropopause within the limits of instrumental accuracy. The difference in speed at the two tropopauses at Caribou was about 5 kt. at 1200 GMT, August 1, and 17 kt. at 1200 GMT, August 2.

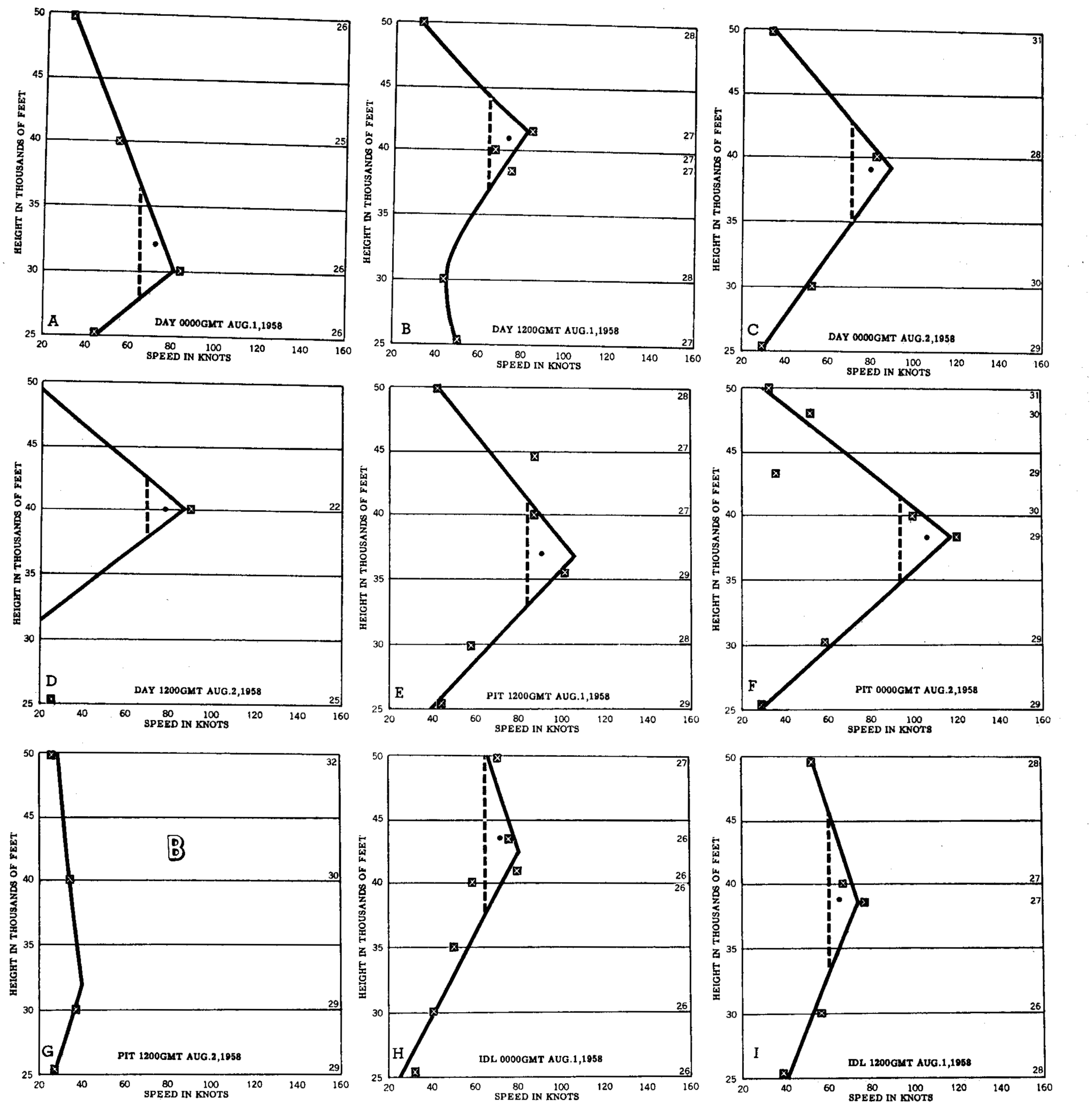
As shown in figure 3B, the height of the peak wind at Idlewild appeared to fluctuate over a layer of 8000 ft. or more from 0600 GMT, August 1 to 0600 GMT, August 2. At the same time, the height of the tropopause showed only minor variations which tended to confirm Endlich's [2] conclusion that the correlation between the height of the peak wind and the tropopause is not too significant for prognostic purposes. Also a comparison of the peak wind (center panel, fig. 3B) and maximum vertical wind shears (lower panel, fig. 3B) at Idlewild showed little or no relationship.

At Flint (fig. 3C), the height of the tropopause and the height of the peak wind appeared almost ideally related. By contrast to this ideal relationship, the vertical wind shears at Flint, though small, were incompatible with the model. Reiter [7] and others have found that the magnitude of negative shears is slightly larger than that of positive ones. During August 1–2, the magnitudes of Flint's vertical wind shears were the reverse of those found in the model.

The relative strength of the positive and negative shears at Dayton (fig. 4A) fluctuated in random fashion during the period and showed little conformity to the jet stream model. Another incompatible feature in Dayton's wind occurred at 1800 GMT, August 1. At this time, the height of the peak wind was reported at 30,000 ft., which was more than 10,000 ft. lower than previous continuity would suggest. As was the case at Idlewild, there appeared to be little or no correlation between the speed of the peak wind at Dayton and the magnitude of the vertical wind shears.

Of all the winds near the core of the jet stream during August 1–2, those at Pittsburgh (fig. 4B) seemed to fit the model best. The magnitude of the negative shears was larger than that of the positive shears in two out of three cases, and the largest shear tended to occur with the strongest wind. The heights of the tropopause exceeded the heights of the peak wind, but the difference between the two values averaged more than 10,000 ft. which seemed quite large when compared to Endlich's and Riehl's models.

Most of the unrepresentative wind reports found near the core of the jet stream were attributed to the inability of wind-measuring equipment to measure accurately strong winds at high altitudes because of low elevation angles. According to U. S. Signal Corps tests [4] on GMD–1 wind equipment, the root-mean-square wind error was found to be about 18 kt. at an altitude of 40,000 ft. and an elevation angle of 6 degrees. Since numerous oscillations of wind speed with height are often found in a complete rawin sounding, an error of 18 kt. at either the top or bottom of a 1000-ft. layer could have introduced a fictitious vertical wind shear of 18 kt./1000 ft. If both the top and bottom were in error, a random relationship could have yielded a



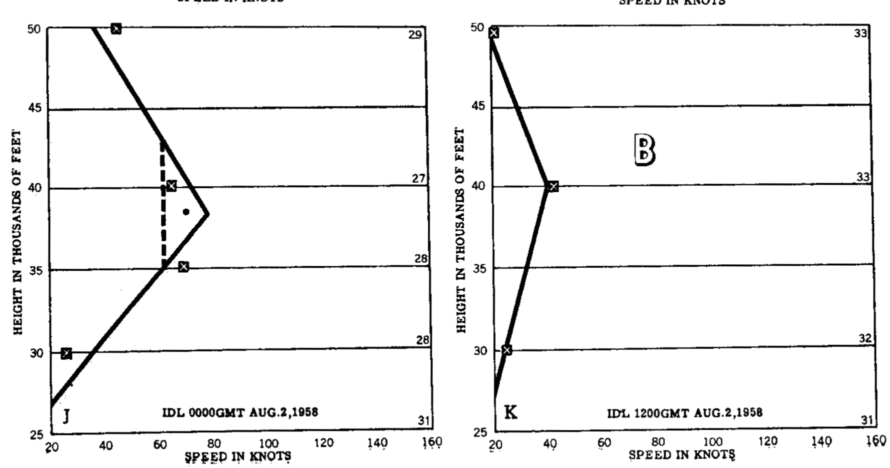


Figure 5.—Smoothed vertical wind profiles along jet stream. "X" reported wind speed, with direction in 10's of degrees at right edge of each section. Dot shows maximum wind (90 percent of peak) at middle of maximum wind layer defined by dashed line connecting points of 80 percent peak wind. Large B = barotropic. DAY = Dayton, Ohio; PIT = Pittsburgh, Pa.; IDL = Idlewild, N. Y.

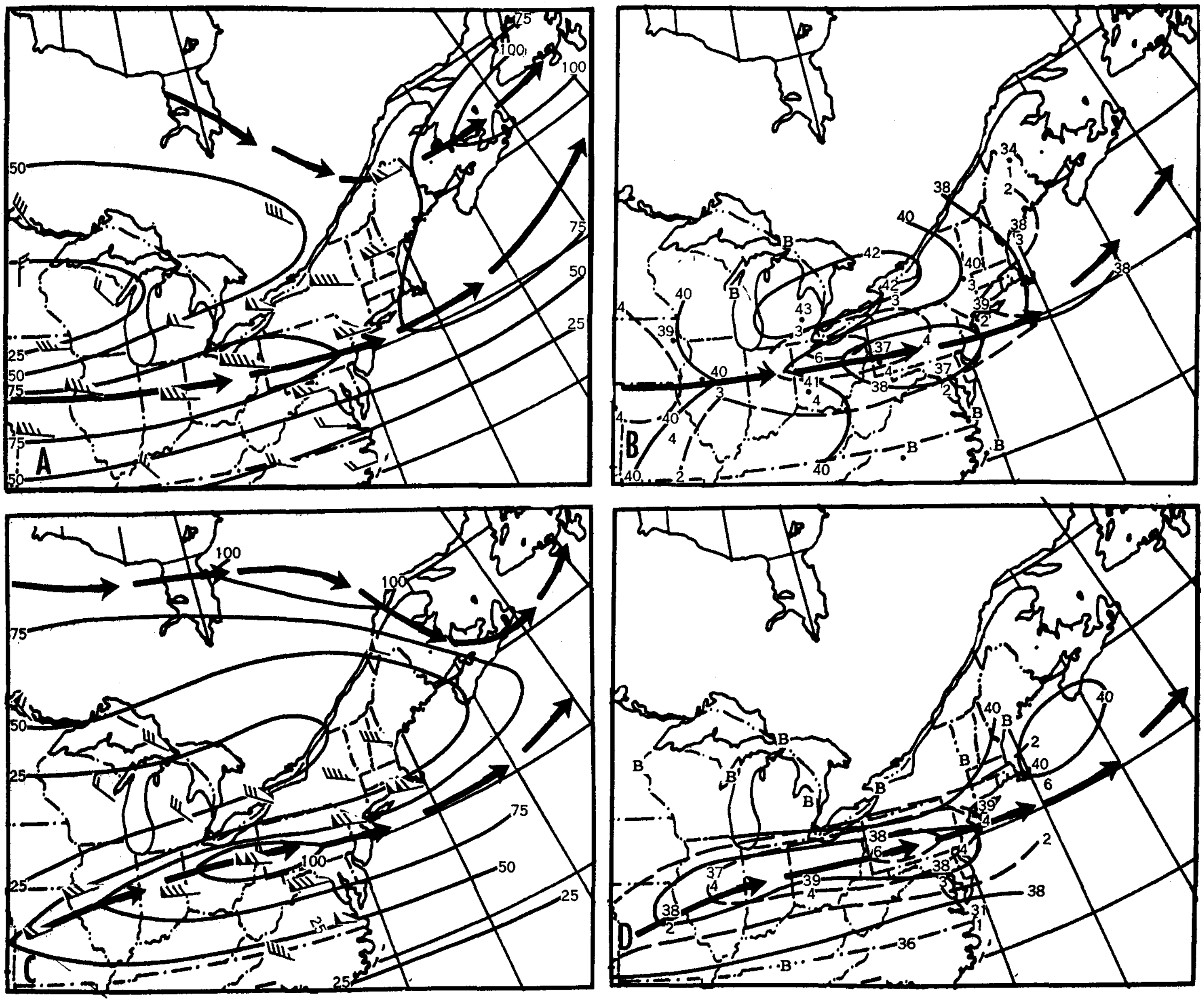


Figure 6.—Vertical wind shear charts which represent the 3-dimensional aspects of a smoothed wind field. Maximum winds and isotachs for (A) 1200 GMT, August 1 and (C) 0000 GMT, August 2 are within 10 percent of the reported peak winds and consistent with the 250-mb. pattern. The height of maximum winds and the averaged vertical wind shears (dashed lines) for (B) 1200 GMT, August 1 and (D) 0000 GMT, August 2 show the topography of the jet stream. Strongest vertical shear values lie north of the jet stream and their 12-hour displacement suggests continuous eastward movement.

maximum fictitious shear of 36 kt./1000 ft. Questionable vertical wind shears similar to these values occurred at both Idlewild and Pittsburgh and were observed near the 40,000-ft. level in conjunction with peak wind speeds of 90 kt. and 119 kt., respectively.

4. ANALYSIS OF VERTICAL WIND SHEAR CHARTS

In the vertical wind shear charts prepared experimentally at NAWAC, the unrepresentative wind features near the core of the jet stream during August 1–2 appeared to be smoothed into a systematic and useful analysis. The procedure of constructing these charts at NAWAC is based on a technique proposed by Reiter [7], with the main difference being that the vertical wind shears are averaged over a 10,000-ft. layer above and below the level of maximum wind. Then isolines are drawn directly for vertical shear values instead of depth of shear layer values as proposed by Reiter. In this manner, the magnitude of average vertical shear can then be used as an incremental term for rapidly estimating wind speeds for almost any altitude between the 500-mb. and 100-mb. levels by subtracting the shear term from the maximum wind value.

Smoothed vertical wind profiles for several stations along the southern jet stream for the period August 1–2 are shown in figure 5 to illustrate the procedure of analysis. In general the profiles were drawn subjectively to conform with the assumption that winds near the jet stream increase and decrease linearly and have one peak velocity. In figure 5 the maximum wind is shown as the point equal to 90 percent of the peak wind, and the depth of the layer of maximum wind is estimated as the distance between points on the profile where the wind is equal to 80 percent of the peak wind. In this manner any wind speed within the layer of maximum wind is assumed to be within 10 percent of the maximum wind.

In figure 5A the smoothing procedure elevated the height of the maximum wind which was placed in the middle of the layer of maximum wind. Another item of interest is shown in figure 5 C, E, and G, in which the presence of an unreported wind was assumed by the straight-line technique. Both of these features obtained from the smoothed wind profile agreed with continuity and seemed quite possible since rawin records are known to contain numerous oscillations in speed which must be interpreted and smoothed before being reported.

Ignoring the unrepresentative wind at 43,000 ft. made the wind profile at Pittsburgh for 0000 GMT, August 2 (fig. 5F) almost classic. The winds varied linearly with height and had one peak velocity. The vertical wind shears appeared to be greater above the peak wind than those below. Vertical wind shears computed directly from rawin data were 9 kt./1000 ft. below and −21 kt./1000 ft. above the peak. When smoothed and averaged over a 10,000-ft. layer above and below the level of maximum wind, the magnitude of the vertical wind shears was reduced to only 6 kt./1000 ft. which was in good agreement with the 5.4 kt./1000 ft. shear computed from the thermal wind equation as previously noted.

The wind profiles at Pittsburgh and Idlewild (fig. 5 G and K) for 1200 GMT, August 2, showed a peak wind of less than 50 kt. Profiles which fell into this category or had a layer of maximum wind greater than 15,000 ft. were classified as "barotropic" (B) since there was little or no vertical wind shear.

Although the extreme vertical wind shears observed during August 1–2 had been rendered imperceptible by the smoothing process, the resulting values presented an organized three-dimensional picture of the wind field near the core of the jet stream when plotted on the vertical wind shear chart (fig. 6). The jet stream position for 1200 GMT, August 1 (fig. 6A) was in essentially the same position as that on the 250-mb. chart (fig. 1B). The isotach pattern near the core of the jet had been altered from that at the 250-mb. level, but then the winds had been computed with a 10 percent tolerance and were representative of a layer about 7000 ft. deep (fig. 5 B, E, I).

The height of the jet stream core (fig. 6B) appeared to undulate from 40,000 ft. at Dayton to 37,000 ft. at Pittsburgh and back again to 39,000 ft. at Idlewild. In relation to the synoptic situation, the lower heights of the jet stream core seemed to be associated with the position of the flat short-wave trough at the 250-mb. level (fig. 1B).

Although there were no extreme vertical shears reported along the southern jet stream at this time, both Dayton and Pittsburgh reported modest and reasonable values of 8 kt./1000 ft. These values were reduced to 4 kt./1000 ft. during the smoothing process. In relation to the surrounding data (fig. 6B) shear values of at least 4 kt./1000 ft. appeared to exist not only along the jet axis in Ohio and Pennsylvania but also about 2° of latitude north of that position and suggested the possibility of at least a 6 kt./1000 ft. shear in an area without data. This pattern was in substantial agreement with the jet stream model.

By 0000 GMT, August 2, the jet stream (fig. 6C) was in about the same position as that 12 hours earlier and was in good agreement with the 250-mb. chart (fig. 1C). The height of the jet core (fig. 6D) varied between 38,000 and 39,000 ft., and followed continuity, with the lower height values appearing to the rear of the flat short-wave trough at the 250-mb. level. Strong horizontal wind shears were observed to the north of the jet maximum, and it was at this time that Pittsburgh reported an extreme vertical shear of 21 kt./1000 ft.

During the smoothing process, this extreme shear at Pittsburgh was reduced to a value of 6 kt./1000 ft.; however, there was now 12-hour continuity for a shear of the lesser magnitude. While this interpolated value of 6 kt./1000 ft. had moved eastward into Pennsylvania, the 4 kt./1000 ft. isoline had progressed to Idlewild. Although the period of continuity had been relatively short, the extreme vertical wind shears near the core of the jet stream during August 1–2 seemed to fit a more usable pattern when smoothed than when allowed to fluctuate in random fashion.

As previously discussed, the procedure of averaging wind speeds and vertical shears appeared to be justified on the

TABLE 2.—*Vertical wind shear (kt./1000 ft.)*

Station	Greatest (+ or −) computed from rawin data	Computed from thermal wind equation	From smoothed profile (20,000-ft. layer)
0000 GMT, August 1			
Idlewild	20	2.5	2.5
Dayton	7	1.5	3.4
Pittsburgh	(1)	2.1	(1)
Flint	4	1.5	1.6
Caribou	−10	1.3	4.7
1200 GMT, August 1			
Idlewild	−3	2.3	4.0
Dayton	8	3.9	1.5
Pittsburgh	−8	5.4	4.1
Flint	5	3.7	2.2
Caribou	−18	.86	1.2
0000 GMT, August 2			
Idlewild	9	2.4	3.5
Dayton	−5	1.8	3.6
Pittsburgh	−21	5.4	6.1
Flint	3	1.5	<1
Caribou	−2	1.5	<1
1200 GMT, August 2			
Idlewild	−2	3.5	1.4
Dayton	9	.81	6
Pittsburgh	2	3.2	<1
Flint	1	.58	<1
Caribou	−3	2.3	1.5

[1] Missing.

basis that wind equipment in current use does not have sufficient accuracy reliably to detect small-scale phenomena in the upper troposphere. The microscale feature of clear air turbulence has been found [8] by numerous jet aircraft penetrations to have an average size of less than 50 mi. in length and less than 2000 ft. in depth. In view of these considerations, it is believed more desirable to locate zones of turbulence with reference to a model rather than to unrepresentative wind data of questionable value.

5. CONCLUSIONS

Most of the extreme vertical wind shears computed from rawin data near the core of the jet stream during August 1–2 showed little or no agreement with the jet stream models, lacked continuity, and appeared unrepresentative when compared to surrounding stations. When the wind data were smoothed to remove the unrepresentative features and the vertical shears were averaged over a 20,000-ft. layer, a pattern emerged on the vertical wind shear chart that indicated some prognostic value in planning wind forecasts for jet aircraft. A comparison of the greatest vertical wind shears computed directly from rawin data with those computed from the thermal wind equation and those obtained from the smoothed wind profile is made in table 2. The largest average vertical wind shears were compatible with the jet stream model and were located north of the jet in conjunction with the strongest horizontal wind shears and temperature gradient. Although some errors were undoubtedly introduced in the smoothing process, they were considered insignificant in the broadscale sense.

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Correspondence

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divergence of the actual flow is replaced by that of the normal, which presumably is in turn a reflection of the large-scale planetary influences mentioned earlier.

Variants of this method, based on recognition that the second term on the right of equation (3) is related to the errors of the barotropic model, have been suggested by Berson [2] and Williams [3].

The relation to Wolff's model may be seen if the basic current is defined to be the sum of the first three harmonics of the actual flow pattern. The analogy becomes more exact if the fictitious changes in the three harmonics are removed at the end of each time-step. Perhaps some difference still remains due to the fact that Wolff's model permits the changes (or tendency) of the three harmonics to be influenced by non-linear interaction with the shorter waves, while this is not permitted when using equation (3).

A method very similar to that of Wolff, but dealing with one-dimensional wave motions, is that proposed by Graham [4].

Another in this family is the present operational model of the Extended Forecast Section, U. S. Weather Bureau, proposed by Namias [5]. Here a "fictitious" wind ($\mathbf{V}_F$) is defined by the equation:

$$\mathbf{V}_F = \mathbf{V} - (\mathbf{V}_B - \mathbf{V}_B^\phi) \tag{4}$$

where $\mathbf{V}$ is the actual wind, $\mathbf{V}_B$ is the basic current at the same point and $\mathbf{V}_B^\phi$ is the latitudinal average, or zonal component of the basic current for the latitude of $\mathbf{V}$. The predictive equation is then obtained by assuming that the absolute vorticity of the fictitious current is conserved:

$$\frac{\partial \zeta_F}{\partial t} = -\mathbf{V}_F \cdot \nabla \eta_F \tag{5}$$